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Advection-Dispersion Interpretation of Tracer Observations in an Aquifer

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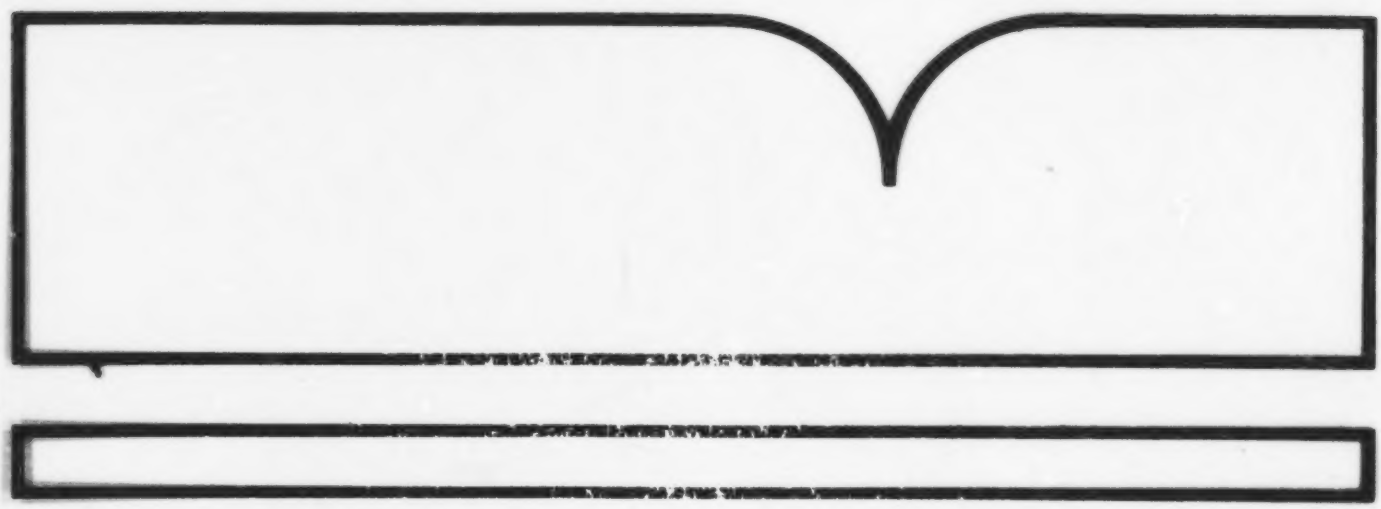

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Advection-Dispersion Interpretation of Tracer Observations in an Aquifer

by Eduard Hoehn[a] and Paul V. Roberts[b]

ABSTRACT

Data from field tracer experiments are interpreted to evaluate the adequacy of an advection-dispersion model for simulating field conditions. A stimulus-response approach based on chemical reactor theory was applied. Two different pulse stimuli at an injection well resulted in responses in two observation wells located at distances of 8 and 17 m, respectively, from the injection well. The wells are perforated over the entire aquifer thickness of 1-2 m. The behavior of the two conservative tracers agreed closely with each other. Breakthrough response curves revealed extended trailing edges (tailing), especially at the nearer well. A two-domain model is employed to extend the simple advection-dispersion equation to account for the observed tailing. Response curves of concentrations were fitted with finite-difference simulations using the two-domain model. The simulations using the two-domain model agreed well with the field observations. The responses at the more distant well were characterized by values of dispersivity equal to or slightly smaller than those at the nearer well.

INTRODUCTION

Using an advanced waste-water treatment facility and a well field operated for artificial recharge in the Palo Alto Baylands (California), investigators at Stanford University have been evaluating whether direct injection of municipal effluents is a reliable and feasible strategy for producing reclaimed water (Roberts *et al.*, 1978). The objective was to provide a framework for characterizing transport of trace organic contaminants in ground water. To complement this work, field experiments using conservative tracers have been carried out to evaluate the adequacy of an advection-dispersion model.

The principles governing contaminant transport in the saturated subsurface environment are reasonably understood in many respects. An approach was adapted from chemical reactor theory for interpreting the responses at observation wells to a stimulus at the injection well (Levenspiel, 1972; Roberts *et al.*, 1980). The theory presented here was used to describe data from a pulse stimulus of nonreactive solutes in one-phase, radially divergent flow in a saturated porous medium. The phenomena of advection and dispersion are emphasized. Extended tailing of the tracer response curves was observed which could not be explained by applying the simplest form of the transport equation (Duffy *et al.*, 1981). This paper evaluates the effectiveness of a two-domain model (Ivanovich and Smith, 1978; Kreft and Zuber, 1979) in simulating the tracer responses.

THEORY

Radial Flow

The hydrodynamic principles governing radial flow from a recharge well have been reviewed by Bear (1972) and by Hoopes and Harleman (1967). Away from the immediate vicinity of a recharge well, flow through a sufficiently permeable porous medium is laminar and governed by Darcy's law

$$q = -K \cdot \frac{\partial h}{\partial r} \qquad (1)$$

where

- q = specific flux, or Darcy velocity (m/h);
- K = hydraulic conductivity (m/h);
- h = hydraulic head (m); and
- r = radial distance (m).

The macroscopic discharge rate q is proportional to the hydraulic gradient $\partial h/\partial r$. At any radius, it is obtained from the continuity equation as

$$q = Q_{IW}/2 \cdot \pi \cdot b \cdot r \qquad (2)$$

where

[a]Department of Applied Earth Sciences, Stanford University, Stanford, California 94305; now at Swiss Federal Institute for Water Resources and Water Pollution Control, EAWAG, CH-8600, Dübendorf, Switzerland.

[b]Department of Civil Engineering, Stanford University, Stanford, California 94305.

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Q_{IW} = injection rate (m^3/h); and

b = aquifer thickness (m).

It is related to the average linear velocity $\bar{v}$ by

$$q = \bar{v} \cdot p \qquad (3)$$

where

p = effective porosity.

Combining equation (1) with equation (2) and integrating gives

$$K = \frac{Q_{IW} \cdot \ln(r_2/r_1)}{2 \cdot \pi \cdot b \cdot (h_1 - h_2)} \qquad (4)$$

According to Bear (1972), the transport equation (advection-dispersion equation) for steady single-phase radial flow in a saturated, homogeneous and isotropic porous medium for a nonreactive solute may be expressed as

$$p \frac{\partial C}{\partial t} = D' \frac{\partial^2 C}{\partial r^2} - q \frac{\partial C}{\partial r} \qquad (5a)$$

or

$$\frac{\partial C}{\partial t} = D \frac{\partial^2 C}{\partial r^2} - \bar{v} \frac{\partial C}{\partial r} \qquad (5b)$$

where

C = dimensionless concentration of solute, explained below;

t = time (h);

D = hydrodynamic dispersion coefficient (m^2/h); and

D = D'/p.

Commonly used initial and boundary conditions of experiments which form a basis for the solution of equation (5) are

(a) for the step stimulus:

$t \leq 0$	$r > r_w$	$C = 0$	(6a)
$t > 0$	$r = r_w$	$C = C_o$	(6b)
$t > 0$	$r = L$	$\partial C/\partial r = 0$	(6c)

where

r_w = well radius (m);

C_o = initial concentration of solute (dimensionless); and

L = distance from injection to observation well (m);

and (b) for the pulse stimulus:

$t \leq 0$	$r > r_w$	$C = 0$	(7a)
$t = 0$	$r > r_w$	$C = C_o \cdot \delta(r)$	(7b)
$t > 0$	$r = L$	$\partial C/\partial r = 0$	(7c)

where $\delta(r)$, the Dirac-function, is described as

$$\left[\int_{-\infty}^{+\infty} \delta(r = 0)\, dr = 1;\ \delta(r \neq 0)\, dr = 0\right].$$

Stimulus-Response-Methodology

The general methodology used here for describing advection and dispersion phenomena is based on a stimulus-response approach (Levenspiel, 1972; Levenspiel and Bischoff, 1963; Roberts *et al.*, 1980), valid in principle for any reactor vessel. The response to a step stimulus is called the $\hat{F}$-response; the response to a pulse stimulus is called the $\hat{C}$-response. The $\hat{C}$-response is the first derivative of the $\hat{F}$-response; the $\hat{F}$-response is the integral of the $\hat{C}$-response. This principle can be utilized to estimate the $\hat{F}$-response from field measurements of the $\hat{C}$-response of a conservative tracer or vice-versa (Levenspiel, 1972; Levenspiel and Bischoff, 1963; Roberts and Valocchi, 1981).

Dimensionless concentrations are defined as ratios of the observed concentration to the input concentration (C/C_o = fractional breakthrough) for the step stimulus; in the case of the pulse stimulus, the dimensionless concentration is expressed relative to a fictitious concentration corresponding to complete mixing of the mass of tracer throughout the pore volume subtended by the observation well, defined below. For either the step or the pulse stimulus, the dimensionless concentrations are plotted as a function of the dimensionless residence time parameter $\theta = t/\bar{\tau}_{H_2O}$, where $\bar{\tau}_{H_2O}$ is the mean residence time of the injection water. For the case of a steady injection in the absence of regional natural flow, fractional breakthrough also can be plotted as a function of cumulative injected volume of water, V_{IW} (Roberts *et al.*, 1980). The mean residence time $\bar{\tau}_{H_2O}$ of the injected water can be estimated from the concentration response of a conservative tracer at an observation well.

A conservative tracer interacts with neither the aquifer nor the pore water and is hence transported at the same velocity as the moving water. It shows a complete breakthrough, i.e. at $t = \infty$, $C(L)/C_o = 1$ for the $\hat{F}$-response; $C(L)/C_o = 0$ for the $\hat{C}$-response. The mean residence time $\bar{\tau}_{H_2O}$ of injected water within the aquifer element is

$$\bar{t}_{H_2O} = \frac{\int_0^\infty (1 - f_{IW})\, dV_{IW}}{Q_{IW}} \qquad (8)$$

and

$$V_p = \int_0^\infty (1 - f_{IW})\, dV_{IW} \qquad (9)$$

is the effective volume of pore water ultimately displaced by injected water in the element, and f_{IW} is the fractional concentration (C/C_o) of injected water in an observation well sample.

In this work, an element volume V_{aq} was calculated as the volume of a cylindrical element corresponding to ideal radial flow and assuming constant average thickness:

$$V_{aq} = \pi \cdot r^2 \cdot b \qquad (10)$$

It is related to effective porosity by

$$V_{aq} = V_p/p \qquad (11)$$

In ideally radial flow in a homogeneous aquifer of constant average thickness, $\bar{t}_{H_2O}$ can be represented from equations (8)-(11) as

$$\bar{t}_{H_2O} = \pi \cdot r^2 \cdot b \cdot p/Q_{IW} \qquad (12)$$

and from equation (11), effective porosity is given as

$$p = \bar{t}_{H_2O} \cdot Q_{IW}/\pi \cdot r^2 \cdot b \qquad (13)$$

The integral in equation (8) can be evaluated from field tracer experiment data. The values of the fractional breakthrough for a conservative solute are plotted against the cumulative volume of injected water. The volume of pore water displaced is estimated by integrating the area above the F-response curve.

Effect of Dispersion

For ideal advective flow (plug flow) without dispersion, a stimulus of a conservative solute is propagated as a sharp wave. Dispersion acts to smear the sharp front, resulting in an S-shaped wave for step stimulus, or a bell-shaped wave for pulse stimulus, with mean residence time equal to $\bar{t}_{H_2O}$. The importance of the dispersion phenomenon can be gauged by the magnitude of the dispersion coefficient D in equation (5). In one-dimensional ground-water hydrology, the dispersion coefficient is frequently found to be approximately proportional to the average linear velocity,

$$D = \alpha_l \cdot \bar{v} \qquad (14a)$$

or

$$D' = \alpha_l \cdot q \qquad (14b)$$

where α_l is known as the longitudinal dispersivity. The contribution of molecular diffusion is considered negligible (Bear, 1972). In most instances, neither laboratory experiment nor theory is of great value in predicting the value of D for an aquifer. The value of D is determined primarily by spatial variation of hydraulic conductivity (Mercado, 1967), and hence must be determined by tracer experiments in the field. A useful measure of the dispersion tendency is the parameter of dimensionless distance, the Peclet number (Pe), defined as Pe = (average linear velocity) × (distance)/ (dispersion coefficient). In radial flow, Pe is given as (Bear, 1972):

$$Pe = r/\alpha_l \qquad (15)$$

For large values of Pe ($Pe > 100$), the responses are nearly symmetric, the relative extent of dispersion is small and the spectrum of the tracer's residence times is relatively narrow; the fractional breakthrough C/C_o is equal to 0.5 when one pore volume of water has been displaced. For small values of Pe ($Pe < 10$), pronounced mixing and a broad spectrum of residence times are evident; the fractional breakthrough corresponding to displacement of one pore volume is appreciably greater than 0.5 (Ogata and Banks, 1961).

Comparison of Different Solutions to the Transport Model

Sauty (1980) compared analytical and numerical solutions of the responses to step and pulse stimuli in both radial and one-dimensional uniform flow ($\bar{v}$ = const.) using dimensionless variables. For the case of a pulse stimulus, he started from the assumption that no exact solutions to the transport equation for radial flow are available. For the response to a pulse stimulus in radially divergent flow, Sauty found the best agreement to his numerical solutions to be the first derivative of the exact analytical solution of Ogata and Banks (1961) for a response to a step stimulus in one-dimensional uniform flow, valid for $Pe > 3$. From Figure 17 of Sauty (1980), it is clear that a pulse stimulus leads to similar response curves when represented in nondimensional form, regardless whether the flow field is radially divergent or uniform.

Limitations of the Transport Model

The above presentation of the transport of nonsorbing solutes in aquifers is premised on the suitability of a diffusion-type transport model. Advection-dispersion models of the simple type

represented by equation (5) have been questioned on various grounds:

Scale Dependence – Field dispersivity values in granular aquifers obtained from results of tracer experiments range from 0.012 to 15.2 m, or up to five orders of magnitude higher than values measured in the laboratory (Pickens and Grisak, 1981). In field studies, the dispersivity has been found to be scale-dependent, contrary to the premise of the basic transport model (Anderson, 1979). Investigators have been compelled to hypothesize scale-dependence of dispersivity to simulate field data. Increases in dispersivity with increasing distances can be rationalized to be caused by the effects of the increasing scale of heterogeneities (Fried, 1975).

Layering – In many situations it must be recognized that an aquifer of interest may consist of two or more distinct layers, each transporting solutes at different rates. It is known that if a point-sampling technique is applied and individual "layers" of the aquifer are sampled, dispersivities tend to be lower than those averaged over the entire aquifer thickness (Pickens and Grisak, 1981). Kreft *et al.* (1974) and Gaillard *et al.* (1977) presented solutions for two-layer and three-layer models. The leading edge of the curve up to about one pore volume of injected water is explained by fast transport through a highly permeable layer. Slow transport through one or two less permeable layers with different dispersivities and porosities was necessary to explain the observed tailing of the response curves. This model, however, was not supported by geological evidence in all cases. Unless the introduction of more parameters is justified from the knowledge of drillers' logs or aquifer samples, such an interpretation has to be accepted with caution. In most cases there is insufficient geological evidence to support the assumption of distinct, continuous layers of different hydraulic conductivity. However, Kreft and Zuber (1979) have pointed out that the domains of different hydraulic conductivity need not be present necessarily as distinct layers; an interconnected system of pores and cracks was postulated to behave similarly to a two-layer system.

Mass Transport – A third possible limitation to the simplest form of the advection-dispersion model [equation (5)] is posed by the possible exchange of solute by diffusion into impermeable domains. In some cases the aquifer may consist of an aggregate of highly permeable, coarse-grained regions on the one hand and domains of fine-grained material having low hydraulic conductivity on the other hand (sometimes referred to as "dead-end" pore space of the porous matrix). If the domains of low hydraulic conductivity possess appreciable pore volume for stagnant water, the assumption of equilibrium is invalid. As an effect of mass transport limitations, diffusion of solute into the region of low hydraulic conductivity must be considered. Bulk flow of the so-called mobile water occurs in the coarse, permeable material, but certain pores may be bypassed. Mixing with such immobile water may occur slowly compared to the time of passage of the front in the mobile domain. Such an incomplete mixing process can lead to the often observed tailing of the response curves (Coats and Smith, 1964; Van Genuchten and Wierenga, 1976).

FIELD STUDY

Tracer Experiments

Field tracer experiments were carried out at the Palo Alto Baylands field site in spring 1978 to evaluate the adequacy of an advection-dispersion model for simulating results connected with artificial recharge. Duffy *et al.* (1981) present the field work in detail. Two different tracer experiments of the pulse-stimulus-type using the radioactive isotopes bromine-82 (as Br^-) and tritium (as tritiated water), respectively, resulted in complete breakthrough responses with steeply rising leading edges and gradually falling trailing edges (tailing). The behavior of the two tracers agreed closely. The tracers were injected through a fully penetrating well (well I2); samples for tracer analysis were withdrawn from fully penetrating observation wells at distances of 7.6 m (well P4) and 16.8 m (well S3) (Roberts *et al.*, 1978, Figure 3).

Aquifer Characteristics

The thickness of the aquifer was 1.5 m at well I2, 2.1 m at well P4 and 1.0 m at well S3. The aquifer was considered to be vertically homogeneous, consisting of a poorly sorted alluvial gravel with a median grain size of 1 mm (Figure 1). The silt-clay content of the aquifer at wells I2 and S3 was approximately 10% and 4%, respectively. The average hydraulic conductivity was estimated from pump tests as 0.1-0.2 m/h at well I2 and 0.2-0.4 m/h at well S3. It is believed that the aquifer gradually becomes more permeable and contains gradually less fine-grained material from the region of the injection well I2 to the region of the observation well S3. An effective porosity of 0.22 was measured in a core sample of well P4. In

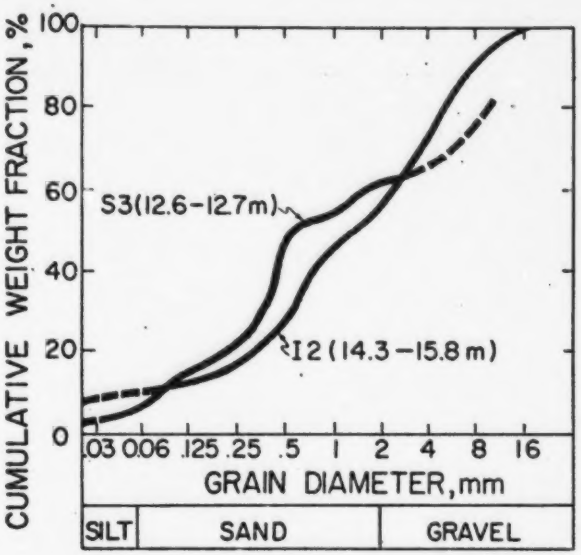

Fig. 1. Grain-size distribution of core samples from the injection well I2 and the observation well S3.

spite of some evidence of upward leakage, the aquifer was regarded as confined in this paper.

Flow Field

Regional natural flow was assumed to be negligible. Water was pumped into the injection well at a steady rate of 2.5 m^3/h inducing artesian conditions at the observation wells. The observation wells were bled continuously: the extraction flow rate at well P4 was 0.34 m^3/h during the bromine experiment and 0.11 m^3/h during the tritium experiment; the flow rate at well S3 was 0.13 m^3/h during both experiments. The steady distribution of hydraulic head (mid-November 1977) at wells I2, P4 and S3 during injection was 10.3, 3.5 and 0.2 m above ground level, respectively (Duffy *et al.*, 1981).

Results and Interpretation

Results of the field experiments can be seen in Table 1. A complete breakthrough occurred at well P4 after about 150 m^3 of water had been injected; for well S3 about 200 m^3 of water was necessary. By integrating above the $\hat{F}$-response curve obtained in turn by integrating the $\hat{C}$-response curve, the effective volumes [equation (9)] of injected pore water for an aquifer element around the injection well I2 up to the observation well P4 were calculated to be about half of the volumes for an element up to well S3 (Figure 2). Calculated mean residence times were found to be about twice as long for well S3 as for well P4. The calculation was made using equation (8), but the extraction

Table 1. Results of the Field Tracer Experiments

Well	Tracer	Pore Volume (m^3)	Mean Residence Time (h)	Average Linear Velocity (m/h)	Effective Porosity §
P4*	^{82}Br	47.5	14.1	0.54	0.15
P4*	^{3}H	54	20.7	0.37	0.17
S3+	^{82}Br	104	ca. 40	0.42	0.08
S3+	^{3}H	117	ca. 45	0.37	0.09

* r = 7.6 m; b_{ave} = 1.8 m.
\+ r = 16.8 m; b_{ave} = 1.5 m.
§ Assuming ideal radial flow [equation (13)].

flow rates were added to the injection flow rate Q_{IW}. The average linear velocity of 0.4-0.5 m/h for flow to both observation wells was nearly the same. This indicates a departure from the case of ideal radial flow, for which the mean residence time should be proportional to the square of distance and the average velocity should be inversely proportional to distance. The lower values of effective porosity (8 to 17%) in Table 1, compared to the measured porosity of 22% also indicate deviation from ideal radial flow. Apparently only one third to two thirds of the cylindrical element was in fact permeated by injected water. This is in accordance with previous interpretations (Roberts *et al.*, 1978, 1980; Charbeneau and Street, 1978) indicating preferential flow in the direction of wells P4 and S3.

The observed lag of the tritium response relative to the bromine response was in part due to the smaller flow rate at the artesian observation well P4 during the tritium experiment. On the other hand, it is known that anions such as Br^- are repulsed by negatively charged clay minerals and hence excluded from interstitial porosity, which tends to increase their apparent transport velocities (Krupp *et al.*, 1972).

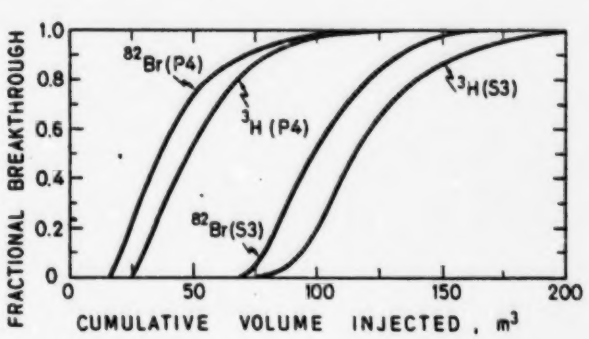

Fig. 2. Tracer breakthrough summation curves ($\hat{F}$-responses).

Table 2. Dispersivity Parameters Using a One-Domain Model

Tracer	Peclet No. (Pe) and Dispersivity (α)			
	P4		S3	
	Pe	α(m)	Pe	α(m)
^{82}Br	10	0.8	50	0.3
^{3}H	20	0.4	50	0.3

CALIBRATION OF THE MODEL

The inverse problem to be solved was to vary the coefficient of dispersivity α until the observed breakthrough response curves were reproduced. The dispersion tendency (in terms of Pe) was first estimated by matching the integration of the $\hat{C}$-response curve ($\hat{F}$-response) with type curve solutions of the simplest transport equation ($\hat{F}$-response in uniform flow) using the Peclet number as a fitting parameter. Such type curves were calculated by Ogata and Banks (1961). The range of dispersivities so obtained is seen in Table 2.

The type curves allowed for neither a representation of the whole of the observed breakthrough responses nor an estimation of the volumes of pore water displaced. A two-domain model extension of the advection-dispersion equation was utilized, according to the proposed approach of Kreft and Zuber (1979), case A, where the flux in the individual domains is considered to be proportional to the volume fraction of the solute passing through the domain. The porosity parameters of the two domains are summed to yield the effective porosity of the aquifer. Given the way the advection-dispersion equation is set up in equation (5a) of this work, the requirements of the two-domain model could be fulfilled by matching the porosity parameter for both domains separately. In ideal radially divergent flow, the porosity parameter p is related to the injection rate Q_{IW} by equation (13).

For the mathematical simulations, a numerical approximation of the advection-dispersion equation [equation (5a)] in ideal radially divergent flow was applied subject to initial and boundary conditions given by equations (7a-c). An improved Crank-Nicholson method which reduces numerical dispersion was used. The method has been described a and verified for the case of uniform flow by Gorelick (1980).

DISCUSSION

As seen in Figures 3 and 4, a two-domain model assuming radial flow simulated reasonably accurately the observed behavior of the tracers. In general, 90% or more of the response curves could be simulated. Values of the transport parameters of the four different sets of simulations are summarized in Table 3. In spite of the observed macroscopic homogeneity of the aquifer in the vertical direction, it appeared that two aquifer domains of different hydraulic conductivities exist which may be discernible only microscopically; water and solutes should be transported through both domains. This may be due to the high nonuniformity of the grain size in the aquifer.

From the steady distribution of hydraulic

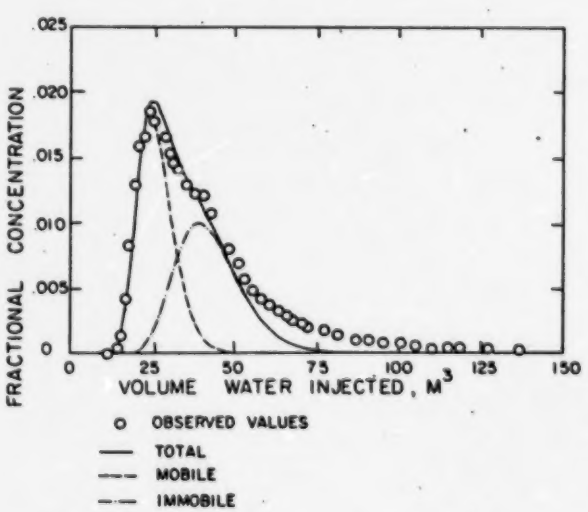

Fig. 3a. Simulation of tracer observations at well P4: ^{82}bromine.

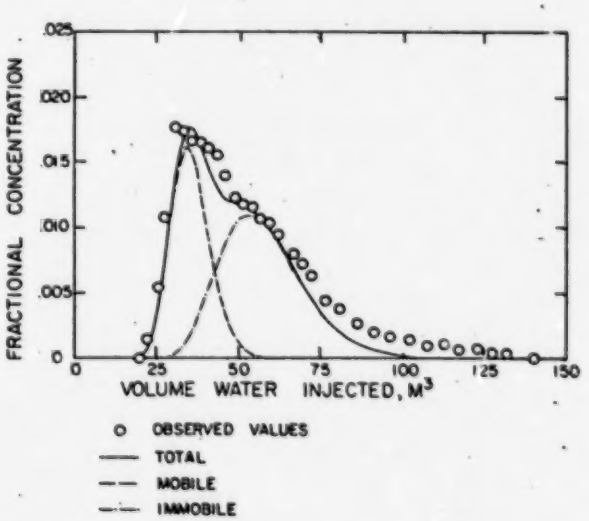

Fig. 3b. Simulation of tracer observations at well P4: tritium.

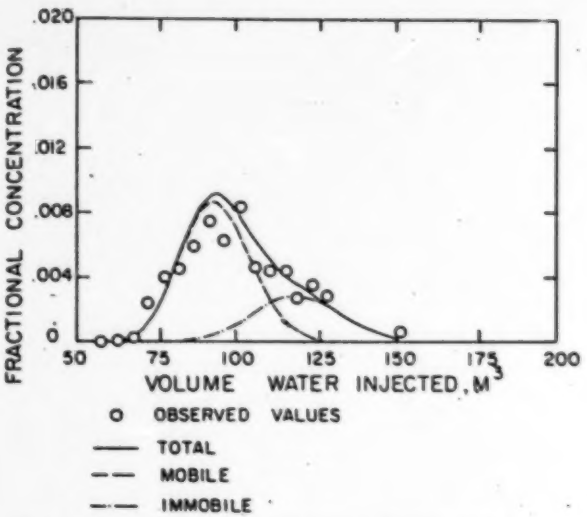

Fig. 4a. Simulation of tracer observations at well S3: ^{82}bromine.

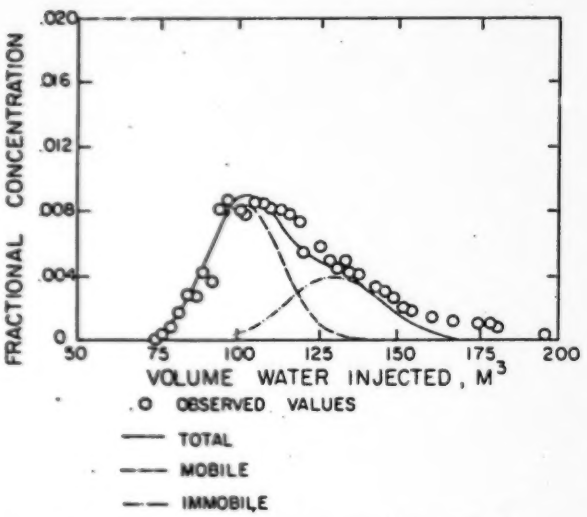

Fig. 4b. Simulation of tracer observations at well S3: tritium.

head around well I2 shown by Charbeneau and Street (1978), it was anticipated that the unknown flow field could be represented approximately by a radially divergent distribution of specific flux. Taking an upwards leakage into account, Charbeneau and Street fitted hydraulic head contours to observed values of August to mid-September 1977. The results of the tracer experiments, however, show a considerable departure from radial flow. The actual flow field probably could be characterized as being intermediate between ideal radial and ideal uniform flow. According to the results of Sauty (1980), the difference in breakthrough response simulations between a model using radial flow and a model using uniform one-dimensional flow is negligible for $Pe > 3$, when expressed in nondimensional variables.

In view of the limitation of the model and the field data, the estimated values for dispersivity must be considered as subject to considerable uncertainty; in this light, differences of less than a factor of two are believed to be insignificant. Dispersivities estimated for the experiment using tritium were slightly smaller than those for bromine, but the differences amount to less than a factor of two and are probably not significant. Likewise, the estimated dispersivities for the two domains of different hydraulic properties differ by less than a factor of two, and hence are considered equivalent. Calculated dispersivities at well S3 were equal to or slightly smaller than at well P4. The tailing at the more distant well S3 was less pronounced than at well P4, and the breakthrough responses at well S3 were nearly symmetrical; the same values for dispersivity were used for the

Table 3. Values of Transport Parameters Estimated Using a Two-Domain Model

Well	Tracer	Type Curve Representing the More Permeable Domain				Type Curve Representing the Less Permeable Domain				Unexplained Volume Fraction
		Volume Fraction (%)	Porosity+	Dispersivity* (m)	Peclet Number	Volume Fraction (%)	Porosity+	Dispersivity* (m)	Peclet Number	(%)
P4	^{82}Br	45	0.04	0.13	60	41	0.07	0.16	47	14
P4	^{3}H	43	0.05	0.08	100	52	0.10	0.14	53	5
S3	^{82}Br	77	0.05	0.09	190	23	0.02	0.09	190	0
S3	^{3}H	64	0.05	0.08	220	31	0.02	0.08	220	5

* Given as $\alpha_l = D/\bar{v}$ [equation (14a)].
\+ Assuming no unexplained volume fraction.

simulations of the leading and the trailing edges of the breakthrough responses at well S3. These slight differences between wells P4 and S3 may be explained by the observed higher hydraulic conductivity and greater uniformity of the aquifer portion near well S3. Moreover, any disturbance of the aquifer during installation of the injection well I2 would affect the response at the nearer observation well (P4) to a greater extent than at the farther one (S3), and hence contribute to a greater apparent dispersivity at well P4.

Constant values of dispersivity at different scales of observation are in accordance with the assumptions of the basic advection-dispersion model (Gelhar and Collins, 1971). However, it has been noted that, in many cases, the apparent longitudinal dispersivity increases with increasing scale (Fried, 1975; Anderson, 1979). The results of this work indicate that dispersivity was not significantly dependent on scale in the range investigated.

Valocchi *et al.* (1981) simulated the breakthrough of injected water at the injection well site I1 in the same recharge area, 1000 m distant from I2. They used a two-dimensional (radial flow) advection-dispersion model and found the optimal value for longitudinal dispersivity to be 1 m. The discrepancies between Valocchi's results and those reported here reflect in part the high spatial variability of hydraulic conductivity in the Palo Alto Baylands area. However, it is recognized that the decomposition of the tracer response into two domains in the present work has the effect of reducing the values of the estimated dispersivity; this arises because part of the variability in residence time that normally would be attributed to dispersion in a simple one-domain model has been accounted for by differences in permeability. In the present case, the apparent dispersivities for the simulation using a single domain (Table 2) are higher by a factor of up to 10 compared to the values obtained for the two-domain model (Table 3).

As an alternative explanation for the tailing phenomena observed in this work, a slow exchange of solute between domains of greater and lesser hydraulic conductivity may be assumed. This case has been treated using mass transfer models by Van Genuchten and Wierenga (1976) and Coats and Smith (1964). Van Genuchten and Wierenga reported that tailing during column experiments using tritium in an unsaturated soil could be explained satisfactorily by accounting for slow mass transfer between mobile and immobile domains. Four physical transport parameters had to be fitted to the model: dispersivity, porosity, a mass transfer coefficient, and the fraction of pore space occupied by mobile fluid. Coats and Smith, who used a similar model in connection with column experiments, pointed out that under field conditions mass transfer into stagnant water volume might be sufficiently rapid compared to the pore-water velocity.

Given a constant mass transfer coefficient, mass transfer will, according to Coats and Smith (1964), be more "instantaneous" with increasing distance. This would be consistent with the observed field data. Further research is necessary to apply a mass transfer model to simulate field observations of breakthrough responses and to compare the results with the two-domain model.

CONCLUSIONS

It is well known that estimating velocity and dispersivity are the most troublesome and expensive aspects of field studies. A detailed knowledge of the local geology and the hydraulic conductivity distribution would be useful. Nevertheless, even in the absence of such knowledge and in cases where the flow field is virtually unknown, tracer experiments are a valuable means for quantifying advection and dispersion during solute transport in ground water. Laboratory studies are of little help for predicting velocity, effective porosity and dispersivity within an aquifer. They must be measured in the field.

In spite of their limitations, advection-dispersion models provided a convenient framework in that it was possible to compare data sets for different tracers and observation points. In its simplest representation, however, the transport equation [equation (5)] did not simulate well the observed field data. A two-domain model accounted better for the observed tailing and the apparent spatial variation of dispersivity, but has the disadvantage of introducing additional parameters.

Numerical solutions of the advection-dispersion equation are a valuable tool for studying field tracer experiments in radial ground-water flow. The amount of field data required to simulate the observations was justified in the light of the overall purpose of gaining insight into the transport behavior of water and solutes during recharge operations.

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Eduard Hoehn received a Diploma in Geology in 1973, and a Ph.D. in Geology in 1979, both from the Swiss Federal Institute of Technology. He was a postdoctoral visitor in Applied Earth Sciences at Stanford University, 1980-1981. He has been a Research Hydrogeologist at the Swiss Federal Institute for Water Resources and Water Pollution Control since 1979. His research interests include ground-water geochemistry, solute transport, and ground-water quality management.

Paul V. Roberts received a B.S. in Chemical Engineering from Princeton University in 1960, a Ph.D. in Chemical Engineering from Cornell University in 1966, and an M.S. in Environmental Engineering from Stanford University in 1971. He has held positions at Stanford Research Institute (1968-1972) and Swiss Federal Institute for Water Resources and Water Pollution Control (1972-1975). He has been at Stanford since 1976, and is currently Associate Professor of Environmental Science and Engineering, Department of Civil Engineering, Stanford University. His research interests include mass transport, trace contaminants, transformations and fates of contaminants in the groundwater environment, phase equilibria, and unit operations of advanced waste treatment.